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Report No. 4
Final Report

Covering the Period 1 June 1962 to 31 May 1963

OBJECTIVE AND DYNAMICAL STUDIES OF TROPICAL WEATHER PHENOMENA

By: R. M. Endlich R. L. Mancuso

Prepared for:

U.S. Army Electronics Research and Development Laboratory
Fort Monmouth, New Jersey

Contract DA 36-039 SC-89092
Project 3A 99-27-025-09-00

STANFORD RESEARCH INSTITUTE

MENLO PARK, CALIFORNIA

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SRI Project 4129

Objective: To carry out research leading to the development of objective methods of analyzing and forecasting tropical weather, and concerning the dynamics of tropical circulations.

Approved by:

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PURPOSE

The purpose of this research is to furnish knowledge that can be used in the development of objective methods of analyzing and forecasting tropical weather and in understanding the dynamics of tropical weather phenomena. It is planned that the objective techniques will be designed for electronic computation in order to gain speed and accuracy, and to reduce personnel requirements in an operational situation. The investigations are divided into the following tasks:

(1) Analyze the three-dimensional structure, the movement, and the surface weather (clouds and precipitation), of selected cases of representative meteorological phenomena, by using conventional methods.

(2) Investigate and, insofar as possible, develop objective analysis techniques applicable to tropical phenomena. In particular, consider automatic computation of a stream function to represent the wind field. Compare the objective analyses with the analyses of Task (1) above.

(3) Utilizing the analyses of Tasks (1) and (2) above, carry out dynamical studies of such topics as the forces predominant in various phenomena, and the conservation of fields of vorticity, divergence and deformation.

ABSTRACT

During the last fifteen years, objective (computer) methods of weather analysis and forecasting have been applied successfully in extra-tropical meteorology, but have not found much application in the tropics. The special difficulties of analysis in the tropics are reviewed. The bulk of this report describes investigations and tests of objective techniques that are applicable to tropical usage.

It is found convenient for several purposes to compute average values of wind components, height, temperature, and humidity in six atmospheric layers. The layer-averaged winds and heights are used to compute a stream function for use in analysis and prediction. The technique of computation fits the value of stream function at each station to a smoothed pattern of the wind components. Wind data within a radius of 9° latitude of the station of interest are considered in the computation, with a heavier weighting given to nearby stations. If the stream function is properly defined, it is convenient to use observed heights as the initial guess. Several Liebmann iterations are made until the computed values stabilize. Applied to wind data arrayed in a square grid, the method of computation is identical to a finite-difference form of a Poisson equation for vorticity; however, the actual computation does not require use of a grid. Instead, computations are made at station locations so that the stream function values may be compared with observed heights. This technique does not require explicit calculations of vorticity and of boundary conditions as required in a stream function obtained from a Poisson equation. Analyses of computed values of stream function are compared with subjective analyses of the same data made independently by Portig, and show excellent agreement.

Computations were also made of divergence, vorticity, deformation, and vertical motion in each of the layers. The average magnitude of divergence (for an average area of approximately 10° lat^2) was 0.9×10^{-5} sec^{-1} while the magnitude of relative vorticity was approximately

twice as large. The dependence of the magnitudes upon scale (area considered in the computation) was also investigated.

An equation is given for computing vertical motion from the change of relative humidity experienced by an air parcel. This equation is believed to be appropriate for use in the tropics, where large spatial and time changes in humidity are observed. An equation is also given for the kinematical advection of quantities such as stream function, humidity, and vorticity. Finally, requirements for further research concerning numerical techniques are summarized.

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LIST OF MATHEMATICAL SYMBOLS

α	A scalar quantity
σ^2	Variance
ϵ	An observational error
r	Correlation coefficient
h	Relative humidity (percent)
T	Temperature (^{o}K)
t	Time
w	Vertical motion (cm sec^{-1})
u_c, v_c	Eastward and northward components of the speed of a conservative quantity
x, y	Eastward and northward coordinates
$\mathbb{V}$	Horizontal wind velocity
$\mathbb{k}$	Vertical unit vector
ψ	Stream function (m^2 sec^{-1})
$\overset{*}{\psi}$	Stream function (m)
χ	Velocity potential (m^2 sec^{-1})
$\overset{*}{\chi}$	Velocity potential (m)
∇^2	Horizontal Laplacian operator
ζ	Relative vorticity ($\mathbb{k} \cdot \nabla \times \mathbb{V}$)
D	Horizontal divergence ($\nabla \cdot \mathbb{V}$)
Z	Height of a pressure surface (m)
f	Coriolis parameter
g	Gravity
μ_1, μ_2	Weighting factors
d	Distance between grid points
β	Northward variation of the Coriolis parameter
lat	Latitude
ω	Vertical motion (dp/dt)

I PUBLICATIONS, LECTURES, REPORTS, AND CONFERENCES

During the course of the contract, conferences pertaining to tropical meteorology were held between various members of the SRI meteorological staff and the following personnel:

Name	Organization	Date and Place
Captain P. Wolff and Staff	U.S. Navy Fleet Numerical Weather Facility	Aug. 29, 1962 Monterey, Calif.
Major General E. Cook	U.S. Army	Sept. 5, 1962 Stanford Res.Inst.
R. Bellucci and J. Walsh	U.S. Army Electronics Research and Development Laboratory	Sept. 12, 1962 Ft.Monroe, Va.
Lt. Col. J. E. Sadler	National Science Foundation Indian Ocean Expedition	Dec. 28, 1962 Stanford Res.Inst.
M. Lowenthal	U.S. Army Electronics Research and Development Laboratory	Jan.28-29, 1963 Stanford Res.Inst.
Dr. W. H. Portig	University of Texas	Mar.18-20, 1963 Stanford Res.Inst.

Two papers were presented by R. M. Endlich and R. L. Mancuso, which reviewed the approach and techniques pursued in this contractual work. They were:

Title	Conference	Date
"Some Applications of Computers in Tropical Analysis"	Fourth Conference on Applied Meteorology	Sept. 14, 1962
"An Objective Stream-Function for Tropical Analysis"	Third Technical Conference on Hurricanes and Tropical Meteorology	June 8, 1963

Also, R. M. Endlich participated in the first and second conferences on Tropical Meteorology, which were sponsored by the U.S. Army Electronic Research and Development Laboratory.

II FACTUAL DATA

A. INTRODUCTION

The historical developments and recent status of tropical meteorology have been described by several authors, including Grimes (1951), Riehl (1954), Palmer (1955), Ramage (1960), and Portig and Gerhardt (1961). In the present report, we will mention only those aspects of tropical meteorology that are pertinent to the particular purpose of this investigation. Stated very briefly, our purpose has been to investigate and develop computer methods of weather analysis and forecasting applicable to tropical regions. The desired methods should be of general application, i.e., not confined to particular geographical regions or weather phenomena. The rationale for this investigation is the following: In extratropical regions, the experience of the last decade has demonstrated that methods of numerical weather prediction produce forecasts of large-scale flow patterns that equal or exceed the skill of comparable subjective forecasts made by experienced meteorologists. The growth of these numerical methods has also lead to the development of objective techniques of analyzing pressure, wind, temperature, and other meteorological quantities. These objective analysis techniques do not require hand plotting of data nor hand drawing of isolines, thus eliminating these laborious and time-consuming operations. From the theoretical standpoint, numerical models, by systematically introducing or excluding various terms in the complex meteorological equations, have added considerably to the understanding of atmospheric processes.

The advantages that are anticipated from the introduction of similar methods into tropical meteorology include increased speed, accuracy, and completeness of analyses and a reduction of reliance upon specially trained individuals. On the other hand, it is not expected that objective methods can be a cure-all for the problems of tropical meteorology. For example, objective methods are handicapped to about the same extent as subjective ones in regions of sparse observations.

Up to the present time, extratropical methods of numerical prediction have not been applied successfully in the tropics (see Jordan, 1955). Factors contributing to this failure are manifold. It is to be hoped that difficulties due to lack of data and of communication facilities will be resolved in the next decade by use of satellite observations of clouds, satellite tracking of constant-level balloons, use of remote reporting stations, etc. However, even in areas of reasonably good data coverage (such as the Caribbean, Africa, India, and parts of the Pacific), major difficulties arise due to the relatively weak circulations, to the weak pressure gradients, to the vanishing of the Coriolis force, to the disturbing effects of local circulations (such as land-sea breezes), and to the errors of measurement. Though the general hydrodynamical equations apply universally, it is not known to what extent simplifying assumptions used in the extratropics can be used in tropical regions. A majority of experience in numerical forecasting has been based upon simplified forms of the vorticity equation wherein vorticity has been obtained by a primary reliance on height observations. The relationship of vorticity and heights is given either by the geostrophic relationship or by the less restrictive balance equation. The extent to which tropical circulations are geostrophic or quasi-geostrophic has not been clearly established and is a matter of current interest. In any case, heights observed in the tropics are not adequate to permit use of the geostrophic or balance equations for describing winds and vorticity. This fact has been known for many years and led to the application of the streamline-isotach method of wind analysis in the tropics. This subjective method of analysis has been generally accepted on the basis of its proven utility. From the standpoint of objective analysis, the streamline-isotach method suffers from the drawbacks that it cannot be easily performed by computer and that the streamlines are not numerical and therefore do not explicitly describe vorticity. The lack of a quantitative stream function has apparently been a major impediment to the application in tropical meteorology of dynamical models based on vorticity considerations. For this reason, a major portion of this investigation has been concerned with formulating, programming, and testing an objective method for obtaining

a stream function. The method that has been developed is described in Sec. II-C.

Other aspects of this study have been concerned with a number of topics, including the following:

(1) The utility of layer-averaged winds, temperature, height, and relative humidity as compared to standard single-level analyses.

(2) The magnitude of errors of measurement and of diurnal changes in the Caribbean.

(3) Magnitudes of divergence, vorticity, deformation, and vertical motion in tropical circulations as a function of scale (i.e., area considered).

(4) Consideration of methods of obtaining heights consistent with observed winds.

(5) Formulation of a method for computing the velocity potential (to represent divergent wind components).

(6) Formulation of kinematical equations for forecasting quasi-persistent scalar quantities, such as stream function or humidity.

In this investigation of tropical meteorology, attention has been confined to a single area of the earth--namely the Caribbean. This area was chosen for several reasons. One advantage of the Caribbean is that observing stations are closer together than in most of the tropics, thus permitting analysis of synoptic-scale patterns. Also, it was believed that actual army operations in the tropics would be accompanied by a network of stations of equal or higher density than the Caribbean network so that objective techniques should be compatible with a relatively high density of stations. Another advantage of the Caribbean is that standard weather observations are available on IBM cards.[1] A final significant

[1] A large supply of IBM cards obtained from the National Weather Records Center was loaned to us by J. R. Gerhardt and W. H. Portig of the University of Texas.

advantage is that the weather patterns in the Caribbean in certain interesting situations had been analyzed at the University of Texas. We have therefore applied the computer programs to the period 5-8 May 1959. Upon completion of our objective analysis of this period, Dr. Portig kindly made his charts available for purposes of comparison (see Sec. II-C).

The figures pertaining to the weather patterns for the period studied are grouped by synoptic hours for ease of reference. All figures appear at the end of the body of this report.

The objective techniques used in the course of the investigation are outlined in block form in Fig. 1. Solid lines indicate completed links and dashed lines indicate links to be developed. It can be seen that the present work is only a beginning toward the development of objective methods of forecasting tropical weather. Further research and experimentation, described briefly in Sec. II-E, will be required to bring objective methods to fruition.

B. GENERAL DISCUSSION OF OBJECTIVE TECHNIQUES

1. Data Processing

In meteorology, it is often convenient to consider the atmosphere as being composed of several layers--such as the surface boundary layer, the trade wind layer, and the tropopause layer. In tropical meteorology, it is common to refer to storms as high-level or low-level depressions, thus implying a two-layer atmosphere. However, any arbitrary division of the atmosphere into layers cannot be expected to delineate the boundaries of various processes, since the importance of such processes varies from case to case. Rather the use of layers is largely a matter of convenience. If several layers are used, the major vertical variations in atmospheric winds, pressure, humidity, etc., can be described. In numerical prediction models used in the extratropics, one to three layers have generally be used; however, some investigators (e.g., Winn-Nielsen, 1961; Charney, 1962) believe that use of additional layers would permit greater accuracy. Besides being compatible with numerical models, the use of layer-averaged quantities tends to suppress

unrepresentative eddies and random errors of measurement. This smoothing property of layer-averages is believed to be of considerable practical importance. For these reasons, the atmosphere from the surface to 60 mb was arbitrarily divided into the following six layers:

Layer I	surface to 850 mb
Layer II	850 to 700 mb
Layer III	700 to 500 mb
Layer IV	500 to 350 mb
Layer V	350 to 200 mb
Layer VI	200 to 60 mb

In each layer, values of u and v wind components, height, temperature, and relative humidity were computed electronically from values tabulated at regular intervals (see Endlich and Clark, 1963) and printed by the computer in mapped form. Examples of such layer-averaged quantities (with isolines drawn by hand) are shown in various figures throughout this report. Layer-averaged heights and temperatures obviously contain observational errors that mask the weak gradients found in the tropics as shown in Figs. 2, 3, and 9-14. These errors cause spurious gradients so that heights and winds appear to be badly out of geostrophic balance. These difficulties with observations are well-known; however, quantitative estimates of errors are not commonly given. Upper bounds on these errors may be obtained from the following considerations, which have been used by Hovermale (1962) to estimate errors in stratospheric data. Consider that the observed value of a scalar quantity (α_o) is composed of the true value (α_T) plus a random error of measurement. Then the observed twelve-hour change (called $\Delta \alpha_o$) is composed of the true twelve-hour change ($\Delta \alpha_T$) and random, uncorrelated, errors (ϵ_1 and ϵ_2) at the beginning and end of the period, i.e.,

$$\Delta \alpha_o = \Delta \alpha_T + \epsilon_2 - \epsilon_1 \tag{1}$$

The variance of these quantities is

$$\sigma^2 (\Delta \alpha_o) = \sigma^2 (\Delta \alpha_T) + 2 \sigma^2 (\epsilon) \tag{2}$$

under the reasonable assumption that (ϵ_1 and ϵ_2) have the same statistical properties. In general, the true variance of atmospheric quantities [σ^2 ($\Delta\, \alpha_T$)] is not accurately known; however, if this quantity is assumed to be zero, Eq. (2) gives an upper bound on the variance of errors. Determined in this way--using rms values of twelve-hour height changes given in Table III (p.18)--the upper-bounds on the variances of height errors in Layers II and V are 85 m^2 and 500 m^2, respectively. Therefore, rms height errors in the upper and lower troposphere are less than 9 m and 23 m, respectively. If we assume (as a rough estimate) that one-third of the total variance is caused by true atmospheric variations, the rms errors in the two layers would be reduced to approximately 7 m and 18 m, respectively. These latter estimates are believed to be fairly reasonable since they are close to those typical of extratropical measurements (e.g., Endlich and Clark, 1963). However, in the tropics where gradients are relatively weak, errors of this magnitude interfere seriously with analyses.

One further point should be mentioned. It is possible that radiosonde instruments calibrated in groups may have errors that have small positive correlations (instead of no correlation, as assumed above). Then the last term in Eq. (2) has the multiplier (1-r) where r is the correlation coefficient, and the upper bounds on errors would tend to be larger than those given. Thus, the assumption that $\sigma^2(\Delta\, \alpha_T) = o$ (which tends to give an overestimate of errors) may be partially compensated by the assumption that r = o (which tends to give an underestimate of errors).

A topic of importance in regard to layer-averaged data concerns the diurnal changes in temperature and height in the tropics. Systematic differences between observations at 00 and 12 GCT might be due to true tidal variations or to instrumental errors that were functions of time of day, perhaps due to radiation errors. For the period under investigation, average temperatures and heights for the Caribbean were computed for each synoptic time. Graphs of these quantities versus time were smoothed to account for synoptic changes. Then deviations of average temperatures and heights at each synoptic hour from the smoothed curves may be attributed to the combination of tidal and instrumental effects. In Layer II, it was found that heights at 00 GCT were approximately 1 m greater than at 12 GCT

and corresponding temperatures were approximately 0.1°C warmer. In Layer V, heights at 00 GCT were approximately 3.5 m greater than at 12 GCT and temperatures were apparently identical. These "diurnal" changes are only small fractions of 12- and 24-hour changes given in Table III and are therefore negligible.

In comparison to layer-averaged height and temperature, layer-averaged relative humidity has several desirable properties. One desirable aspect is the smoothness and regularity of the patterns as shown in a typical case in Fig. 4. The patterns also showed reasonable persistence from one synoptic hour to the next. Close to the oceanic surface (i.e., in Layer I) relative humidity was quite uniform; however, at higher levels large spatial variations in humidity were evidently produced by synoptic-scale vertical motions acting upon the pre-existing humidity distributions. Instrumental errors in measuring relative humidity, which are on the order of ±5% (Handbook of Geophysics, 1960) are only a small fraction of the observed gradients. It appears that relative humidity may be used to estimate vertical motions in a manner analogous to the adiabatic method (which is based upon temperature changes of parcels and therefore cannot be expected to give reliable results in the tropics). The relative humidity equation, which also assumes adiabatic motions, is derived in Appendix A along the lines given by Smagorinsky (1960). It has the following form:

$$w = 2.3\ (dh/dt)\ (T^2/h) \qquad (3)$$

where w has the units cm sec^{-1}, h is relative humidity in percent, t is time in seconds, and T is in °K. Experiments have not yet been carried out in the use of this equation; however, computer programming is expected to be feasible. It appears that this equation may be used to determine w independently of the kinematic method used in Sec. II-D. Presumably, the best estimate of w would be a combination of values obtained from several different methods.

2. Objective Analysis and Forecasting

The layer-averaged heights and wind components have been used for two purposes consistent with accepted practices of tropical analysis.

The first purpose is the objective computation of a stream function. The second use is in the kinematical computation of divergence, vorticity, deformation, vertical motion, geostrophic winds, and geostrophic departures using a formulation of Endlich and Clark (1963). The results of these computations are discussed in Sec. II-D.

We have not used a square grid in either stream function or kinematical computations. Instead, the stream function has been computed at the station locations and kinematical quantities have been computed at the centers of triangles each defined by three adjacent stations. The computations were made in this way for convenience, and to avoid introducing the uncertainties inherent in grid-point analyses of meteorological quantities. It is expected that grid-point analyses of the computed quantities will have certain advantages in further work, and will be introduced at the appropriate stage.

Another topic that has been considered is an objective method of advection of quasi-conservative quantities such as stream function, vorticity, humidity, etc. It is often observed in the tropics that weather systems move fairly regularly. Therefore, equations analagous to Petterssen's rules have been written and appear to provide a means of objectively forecasting those weather systems which move steadily and without large, sudden changes in intensity. The equations are derived in Appendix B and have the simple form

$$
\begin{aligned}
u_c &= -(\partial\alpha/\partial t)\ (\partial\alpha/\partial x)\ [(\partial\alpha/\partial x)^2 + (\partial\alpha/\partial y)^2]^{-1} \\
v_c &= -(\partial\alpha/\partial t)\ (\partial\alpha/\partial y)\ [(\partial\alpha/\partial x)^2 + (\partial\alpha/\partial y)^2]^{-1}
\end{aligned}
\qquad (4)
$$

where u_c and v_c are the eastward and northward components of speed of the conservative quantity α. To use these equations, it will be necessary to develop a computer method of iterative solution. It is expected that this can be done by further research and that forecasts can be obtained rapidly by this technique. Such a technique would be limited in that it would not introduce the dynamics of the weather phenomena. Nevertheless,

it is pertinent to recall the study of Lavoie and Wiederanders (1960) who found that a simple 50-50 combination of persistence and climatology produced forecasts better than the subjective products of tropical meteorologists. The combination of persistence and climatology may be expected to suppress anomalies (e.g., vortices, waves, etc.) by returning unusual values towards normal. It would appear preferable to move the anomalies forward with the speed they have recently exhibited. It is hoped that this can be accomplished by use of the kinematical formulas given above.

The development of dynamical models for forecasting tropical circulations remains as a long-range goal. Due to the multitude of the practical and theoretical problems mentioned above, it was not expected that it would be possible to conduct dynamical experiments during the period of this contract. However, when the proper foundation has been laid, the extensive numerical experiments that have been carried out in extratropical meteorology can be used as a guide for the rapid development of tropical prediction models.

C. THE DIRECT METHOD OF STREAM-FUNCTION COMPUTATION

It was mentioned earlier that two of the main difficulties in applying conventional numerical analysis and prediction in the tropics are the apparent lack of geostrophic flow and the interference caused by height errors. Tropical analysts rely mainly on wind data, which is analyzed subjectively in the form of streamlines and isotachs. However, the stream lines (drawn everywhere tangent to observed wind vectors) do not have numerical values nor is their spacing inversely proportional to wind speed, as is true for the contours of mid-latitudes. To remedy these serious difficulties, it is desirable to develop a computer method for obtaining a quantitative stream function. There are several possible approaches to this goal. The most obvious approach would be to use the geostrophic or balance equations as done routinely in mid-latitudes (Cressman, 1959). However, use of these equantions is not feasible because of the two difficulties mentioned above. A second possibility is to determine the vorticity of the observed winds.

Then, if proper boundary conditions are determined, the stream function can be obtained by solution of a Poisson equation. This method has been used in mid-latitudes by Brown and Neilon (1961), Charney (1962), and others. A third possibility is to determine a stream function from the observed winds without the necessity of explicitly calculating vorticity and without requiring a separate computation of boundary conditions. The method that has been developed for doing this task is described below. It is similar in concept to a technique of hand analysis used by Scott (1958).

From the well-known Helmholtz equation, the horizontal wind vector $\mathbb{V}$ may be represented as the sum of two components, the first being non-divergent and the second being irrotational

$$\mathbb{V} = (\mathbb{k} \times \nabla\psi) + \nabla\chi = \mathbb{V}_{\psi} + \mathbb{V}_{\chi} \tag{5}$$

The vorticity and divergence of $\mathbb{V}$ are then given by the two Poisson equations $\zeta = \nabla^2\psi$ and $D = \nabla^2\chi$. We wish to obtain a stream function (ψ) that represents the components of translation, vorticity and deformation, while excluding divergence which is to be represented by the velocity potential (χ). It is known from empirical investigations that divergent components are relatively small. We will assume for the time being that they can be suppressed by an areal averaging process performed during the stream function computation. From Eq. (5),

$$u_{\psi} = -(\partial\psi/\partial y), \qquad v_{\psi} = (\partial\psi/\partial x) \tag{6}$$

Also, we may write the differential of ψ between two points separated by increments δx and δy as

$$\delta\psi = (\partial\psi/\partial x)\ \delta x + (\partial\psi/\partial y)\ \delta y = v_{\psi}\ \delta x - u_{\psi}\ \delta y \tag{7}$$

These relationships are sufficient for the computation of ψ which can then be used in well-known prediction equations. However, in this formulation, ψ has the units $m^2\ sec^{-1}$ and therefore is not directly comparable to any other meteorological quantity. For purposes of

comparing values of stream function and observed heights, it is convenient to define

$$\overset{*}{\psi} = (f/g)\ \psi \qquad (8)$$

where f is the Coriolis parameter[2] and g is gravity, so that $\overset{*}{\psi}$ has the units of height (e.g., meters). The differential in $\overset{*}{\psi}$ between a station o and a nearby station 1 is $\delta\overset{*}{\psi} = \overset{*}{\psi}_o - \overset{*}{\psi}_1$, or by using Eq. (7) with the average wind components taken as $(u_o + u_1)/2$ and $(v_o + v_1)/2$ it is

$$\overset{*}{\psi}_o = \overset{*}{\psi}_1 + \frac{(f_o + f_1)}{2g}\left[\frac{(v_o + v_1)\ (x_o - x_1)}{2} - \frac{(u_o + u_1)\ (y_o - y_1)}{2}\right] \qquad (9)$$

where subscripts denote values at the two stations. Let us assume that we have an estimate of $\overset{*}{\psi}_1$ and that the wind components are known. Then Eq. (9) can be solved to give $\overset{*}{\psi}_o$. If station o has n neighbors, each of them can be used to give an estimate of $\overset{*}{\psi}_o$. Presumably, the best estimate of $\overset{*}{\psi}_o$ will be an average of the n separate estimates, i.e.,

$$\overset{*}{\psi}_o = \frac{1}{n}\sum_{i=1}^{n}\overset{*}{\psi}_i + \frac{1}{n}\sum_{i=1}^{n}\delta\ \overset{*}{\psi}_i \qquad (10)$$

An illustration of the computation of $\overset{*}{\psi}_o$ (the value of $\overset{*}{\psi}$ at an arbitrary station) is shown in Fig. 7. The estimates of $\overset{*}{\psi}$ at the four nearby stations are listed. Using the value of $\overset{*}{\psi}_1$ (950) and considering the winds at stations o and 1, we obtain an estimate of $\overset{*}{\psi}_o$ of 940. Similarly, we obtain the three other estimates shown. The average value of the four estimates is 948. This value is then retained (as in a Liebmann iteration) while attention is focused on the next station. All stations are considered in an arbitrary order. For convenience, the original estimates of $\overset{*}{\psi}$ are taken equal to reported values of height. Moreover, two estimates of $\overset{*}{\psi}_o$ are obtained. The first estimate considers all

[2]Since f approaches zero as the equator is approached, it is desirable to avoid the use of f when this becomes critical. This can be done by setting $\mathbf{V} = \mathbf{k} \times \nabla\psi$, by letting f be a constant, or by placing a fictitious lower limit on f.

stations lying within a radius of 300 miles (5^{o} latitude) of the station of interest while the second estimate considers stations lying between 300 and 550 miles away. These two estimates are then combined as

$$\overset{*}{\psi}_o = \mu_1 (\overset{*}{\psi}_o)_1 + \mu_2 (\overset{*}{\psi}_o)_2 \qquad (11)$$

where μ_1 and μ_2 are weighting factors, presently set at 0.8 and 0.2. Thus the observations are weighted more or less inversely with distance from the point of interest, as is done in mid-latitudes. The inclusion of winds over a fairly large area also has the characteristic of tending to suppress the smaller-scale divergent wind components. Since observed heights are used as the initial values of $\overset{*}{\psi}_i$, the average value of stream function for the entire region remains within a few meters of the average value of the observed heights. However, the rms value of the individual differences between height and stream function was 12 meters in Layer II, and 29 meters in Layer V. A block diagram of the computational procedure is shown in Fig. 8. The effect of the first scan of the field of observing stations is to strongly modify the reported heights that are in poorest agreement with the winds. This effect is illustrated in Figs. 13 and 14. In the upper part of Fig. 13, the dashed lines are drawn to heights. Lines drawn to the values of $\overset{*}{\psi}$ after one scan are shown by the dashed lines in the lower portion. Obviously, much spurious information has been removed. Several scans of the field are made until nearly convergent values of $\overset{*}{\psi}$ are obtained (shown by the solid lines in the lower portion of Figs. 13 and 14). In Table I, the values are shown at each station after each scan. It can be seen that the major changes occurred during the first few scans. In order to test the influence of the first guess on the computation, a trial was made wherein the initial guess was taken as a constant value at each station. After several more scans than are usually required, the same gradients were obtained as when reported heights were used as the initial guess. Thus, the final result is not sensitive to the initial values. However, if one were to make only one or two scans, one would obtain a height field smoothed by use of wind data. Further scans tend to eliminate the heights (except their average value as mentioned earlier) in favor of the winds.

Table I

VALUES OF STREAM FUNCTION *VERSUS* SCAN NUMBER, SHOWING THE CONVERGENCE OF THE COMPUTATIONAL PROCEDURE. VALUES ARE IN METERS, FIRST DIGIT OMITTED IN LAYER II AND FIRST TWO DIGITS OMITTED IN LAYER V

STATION NO.	SCAN NO. (LAYER II)							SCAN NO. (LAYER V)							
	0*	1	2	3	4	5	6	0*	1	2	3	4	5	6	7
250	370	358	355	355	357	358	358	360	350	369	379	386	389	391	391
251	345	361	358	357	358	360	360	285	329	343	356	364	369	371	372
240	380	371	375	373	375	376	377	322	280	308	318	329	334	337	338
232	376	378	381	382	383	384	384	289	295	311	320	325	328	329	329
221	Msg	377	375	380	382	382	382	Msg	292	292	303	306	307	308	308
206	376	365	368	372	373	373	373	289	264	270	277	278	278	278	277
211	375	365	371	373	373	374	374	269	277	284	287	287	287	287	286
794	357	361	364	367	367	367	368	264	258	268	271	272	272	271	271
202	345	351	357	357	357	358	358	273	273	279	279	279	279	278	278
063	343	348	352	352	352	353	353	256	260	266	266	266	265	264	264
076	338	341	342	341	342	342	342	236	261	264	262	262	261	260	260
089	334	339	335	335	336	336	336	248	271	265	266	265	264	263	262
325	330	352	352	354	354	354	355	286	302	297	299	298	298	297	296
644	357	360	362	363	364	365	365	340	347	356	354	356	356	356	356
692	Msg	362	360	360	362	363	364	Msg	408	426	430	436	438	439	439
501	323	338	334	337	338	339	340	325	345	328	326	324	323	322	322
383	343	334	339	338	339	340	340	334	315	315	310	309	308	308	307
355	346	338	337	338	338	339	339	325	294	293	292	291	290	289	289
397	324	329	330	330	331	331	331	327	313	311	309	308	307	307	306
367	339	332	331	332	332	333	333	293	298	295	295	293	292	291	291
118	355	336	330	331	331	331	331	299	287	286	285	283	282	282	281
467	340	331	329	329	328	328	328	319	318	320	316	315	314	314	313
525	330	332	330	329	328	328	327	342	351	346	344	343	342	341	340
866	337	331	330	328	328	327	327	386	370	364	364	363	362	361	360
897	320	324	321	320	319	319	319	392	383	380	381	379	378	377	377
001	325	311	316	319	320	321	321	391	357	349	344	343	342	341	341
805	294	305	311	314	315	315	316	369	371	360	357	356	355	354	354
988	323	318	314	313	313	312	312	391	358	371	365	365	364	363	363
967	318	311	308	308	307	307	307	321	381	380	378	378	377	376	375

* These values are reported heights taken as the initial-guess field.

Further insight into the workings of the stream function procedure may be gained by applying Eq. (9) to a hypothetical group of stations arranged in a square network--Station o is at the center, Station 1 lies a distance d to the east, Station 2 lies similarly to the north, 3 to the west, and 4 to the south. This equation is used for relating Station o with each of the four neighbors (i = 1, 2, 3, 4). For example, if i = 1 and f is considered constant

$$\overset{*}{\psi}_o = \overset{*}{\psi}_1 + \frac{f}{2g} (v_o + v_1) (- d) \tag{12}$$

The four relations for i = 1, 2, 3, 4 are then summed, giving

$$4\overset{*}{\psi}_o = \sum_{i=1}^{4} \overset{*}{\psi}_i - \frac{fd^2}{g} \left[\frac{(v_1 - v_3)}{2d} - \frac{(u_2 - u_4)}{2d} \right] \tag{13}$$

The quantity in brackets is a finite-difference expression for vorticity (obtained from wind components) and $\sum \overset{*}{\psi}_i - 4 \overset{*}{\psi}_o$ is a finite-difference form of a Laplacian of the stream function. Viewed in this way, the finite difference form of the stream function procedure becomes identical to the usual finite difference form of Poisson's equation. Therefore, when the stream function procedure scans over a dense, evenly spaced network, it will duplicate the solution of the Poisson equation. On the other hand, the solution will depend on Eq. (9) and on the arrangement of stations in a thinly covered region or at regional boundaries.

As mentioned earlier, we have chosen to concentrate our attention on $\overset{*}{\psi}$ rather than ψ, since the former values can be compared directly with height data, and also permit analyses to be continued from mid-latitudes into the tropics. However, in the former case, one should note that the vorticity is

$$\zeta = f^{-1} (g \nabla^2 \overset{*}{\psi} + \beta u) \tag{14}$$

where $\beta = \partial f/\partial y$. Similarly, the divergence is

$$D = f^{-1} (g \nabla^2 \overset{*}{\chi} - \beta v) \tag{15}$$

Thus, the stream function field $\overset{*}{\psi}$ contains the divergence $-f^{-1}(\beta v)$, which is on the order of 10^{-6} sec^{-1}. In the context of numerical forecasting, it may be simpler to utilize ψ rather than $\overset{*}{\psi}$. Either or both quantities can be obtained by a simple option in the computer program.

The direct stream function computation described above appears to have several advantages over other possible formulations. The general approach is simple in principle. The use of explicit boundary conditions is avoided. It is not necessary to transform to a regularly spaced grid. The method does not seem to be subject to a limitation on the computational stability (such as the ellipticity condition $\zeta_g > - f/2$, which applies to the balance equation). Also, the stream function procedure is fast, fairly insensitive to data distribution, operates quite well when data are missing, and could easily incorporate aircraft, constant-level balloon, or satellite wind data.

At present, total machine time for computing $\overset{*}{\psi}$ in each of six layers for the Caribbean and printing out the computed values in mapped form is about eight minutes using an ALGOL program with the Burroughs 220 computer. Various characteristics of the Burroughs machine are compared in Table II with other computers commonly used in meteorological applications.

TABLE II - Comparison of Computers

Computer	Add Time	Core Storage	Word Size	Algebraic Compiler
Burroughs 220	200 µsec	2-10K	10-decimal	ALGOL
IBM 709	24 µsec	4-36K	36-binary	FORTRAN
Contral Data 1604	4.8 µsec	8-32K	48-binary	FORTRAN
IBM 7090	4.4 µsec	32K	36-binary	FORTRAN
Burroughs B5000	3 µsec	4-32K	48-binary	ALGOL

As can be seen, the 220 is--relative to the more advanced computers--a slow and small computer, so that the same computations could be performed much more rapidly using one of the newer machines.

Examples of the stream function ($\overset{*}{\psi}$) fields are shown in Figs. 2, 3, 9, 10, 11, 12, 13, and 14 for Layers II and V. In addition to the stream function, the actual height contours and wind vectors are also portrayed. As mentioned above, one of the advantages of a stream function having the units of meters is that it permits comparison with subjective contour analyses. Dr. Wilfried H. Portig and associates of the University of Texas, who have considerable experience in tropical meteorology, have prepared contour maps for the same period. It is our understanding that Dr. Portig's technique is based primarily on heights corrected by use of thickness charts and time continuity, and to a lesser degree upon winds. His analyses for the 775 mb and 250 mb levels--in close but not identical correspondence to the Layers II and V--are illustrated in Figs. 13 and 14. As can be seen by comparing the solid lines in the upper and lower portions of the figures, there is excellent agreement, while both are significantly in disagreement with the observed heights. Other comparisons of the two methods of analysis have given the same impression. The discrepancies noted can be attributed to the differences between layer and constant-level winds or to the subjectivity involved in both analyses over areas of sparse data.

In evaluating the stream function, it was believed of importance to investigate a quality generally sought in meteorological quantities--persistence--that is, that the quantity remain unchanged for a certain period of time and thus (to a degree) serve as a forecast. In order to investigate the persistence of the stream function and of reported heights, the rms changes and correlations over various time intervals have been computed. The correlation coefficient is particularly revealing because it eliminates consideration of the variations of the mean fields. The results for Layers II and V, derived from the eight synoptic times under investigation, are tabulated in Table III. It is quite apparent from this table that the stream function possessed considerably greater persistence (correlation coefficients between 0.72 and 0.96) than did

TABLE III

COMPARISON OF HEIGHT (Z) VERSUS STREAM FUNCTION ($\overset{*}{\psi}$) FOR 5-8 MAY 1959
(30 Caribbean Stations)

LAYER	PERIOD (hours)	ROOT-MEAN-SQUARE CHANGE (meters)		CORRELATION	
		Z	$\overset{*}{\psi}$	Z	$\overset{*}{\psi}$
II (850-700 mb)	12	13	10	0.80	0.92
	24	16	16	0.81	0.87
	36	23	24	0.75	0.80
	48	30	33	0.72	0.72
V (350-200 mb)	12	32	19	0.73	0.96
	24	40	27	0.86	0.96
	36	51	35	0.79	0.94
	48	55	43	0.80	0.92

the height (correlation coefficients between 0.72 to 0.86). This difference in persistence was particularly noticeable in Layer V.

It is possible that a slightly modified form of the stream function technique could be used to calculate the velocity potential χ. In order to do this, it would be necessary to determine $\mathbf{V}_\chi$ at each station as the difference between $\mathbf{V}$ (the observed wind) and $\mathbf{V}_\psi$, where the latter is determined from the stream function procedure [see Eq. (5)]. Then, analogously to Eq. (7), we can write a differential of χ between two points as:

$$\delta\chi = (\partial\chi/\partial x)\ \delta x + (\partial\chi/\partial y)\ \delta y = u_\chi\ \delta x + v_\chi\ \delta y. \tag{16}$$

Thus, with u_χ replacing v_ψ and v_χ replacing $-u_\psi$ in Eq. (7), the same formal computational procedures (eqs. 9, 10, and 11) can be used. However, an initial estimate of $\overset{*}{\chi}$ at each station is required and no appropriate estimate (analogous to the use of heights in determining $\overset{*}{\psi}$) is available. Thus, it would probably be necessary to attempt to generate the values of $\overset{*}{\chi}$ using a constant value at each station as the initial estimate. This procedure has not been tested, but appears to be a possibility worthy of investigation.

D. KINEMATICAL COMPUTATIONS

Investigations of divergence or vertical motion in the tropics have been made by Landers (1955), Ballif (1958), Rex (1958), and Sinclair (1958). We have also computed the magnitudes of these quantities, as well as vorticity and deformation, for the period under investigation. The method of computation utilizes the layer-averaged winds and considers three stations at a time; however, the procedure is identical for each triangle (i.e., independent of its size and shape). The values of the quantities are printed for analysis at the centers of the triangles. Examples of the fields of relative vorticity, divergence, and vertical motion are shown in Figs. 4, 5, 6, and 15. In spite of the "noise" produced by random observational errors, the fields are reasonably smooth. (As discussed later, further smoothing could be performed objectively

to suppress the smaller features.) In preference to showing a large number of individual charts, we will present a summary of the magnitudes of the various quantities (Table IV). The average magnitude of divergence was 0.9×10^{-5} sec^{-1} for an average area of approximately 10^{o} lat^{2}. This value may be compared with values of 1.5×10^{-5} sec^{-1} given by Landers for the tropical Pacific and a value of approximately 0.5×10^{-5} given by Ballif for the same region, both for areas of 4^{o} lat^{2}. The average magnitude of vorticity in our study was 1.7×10^{-5} sec^{-1}, almost double the magnitude of divergence. In contrast, Baliff found that the two quantities were of nearly equal magnitude. Due to the differences in these estimates, it is probably desirable to make further computations in various regions and under various weather patterns. Values in parentheses in Table IV were obtained from previous computations by the same techniques over the United States for an average area of approximately 8^{o} lat^{2} (using data for the months of January and July). Due to the effects of errors discussed below, all of these computed magnitudes may be expected to be larger than the true magnitudes. Table IV indicates that in Layer II the tropical values are somewhat smaller than the mid-latitude values (in parentheses), as one would expect; however, in Layer V, there is not much difference. Therefore, one may suspect that the upper-air flow during the period of investigation was unusual, and indeed consisted of a sharp trough and unusually strong winds. In order to compare the magnitudes of the various quantities as computed in this study, we may assign vorticity the arbitrary value 1. Then divergence, stretching deformation and shearing deformation have the relative values 0.5, 0.9, and 0.7 respectively. These relative values appear to be about the same in the tropics as in mid-latitudes for the scales of motion considered. It is also of interest to note that the smallest magnitude of divergence was found in Layer III (centered at 600 mbs). This fact suggests that the concept of a level of least-divergence may be as valid in the tropics as it is in mid-latitudes.

Table IV

AVERAGE MAGNITUDES OF VORTICITY, DIVERGENCE, DEFORMATION AND VERTICAL MOTION FOR 5-8 MAY 1959 IN THE CARIBBEAN

LAYER	VORTICITY	DIVERGENCE	DEFORMATION (STRETCHING)	DEFORMATION (SHEARING)	VERTICAL MOTION†
I	1.0 x 10^{-5} sec^{-1}	0.7 x 10^{-5} sec^{-1}	0.9 x 10^{-5} sec^{-1}	0.9 x 10^{-5} sec^{-1}	1.2 cm sec^{-1}
II	1.0 (2.0)*	0.8 (1.2)	0.9 (1.6)	0.8 (1.5)	2.1
III	1.0	0.6	1.0	0.8	3.5
IV	1.6	0.9	1.6	1.1	5.2
V	3.2 (3.5)	1.5 (1.6)	2.7 (2.6)	2.0 (2.5)	11.5
VI	2.3	1.0	1.8	1.5	--
I-VI	1.7	0.9	1.5	1.2	

* Numbers in parentheses are comparable values for the United States.

† At the top of the layer in question.

A glance at the Caribbean map shows that the observing stations are rather unevenly spaced. Therefore, the triangles used in the kinematical computations vary in area from 22,000 km^2 (1.8^o lat^2) to 920,000 km^2 (75^o lat^2). We may therefore ask how the magnitudes of divergence and vorticity change with scale (i.e., size of the area considered). The correlation of instantaneous values of divergence or vorticity with area is not high. Nevertheless, magnitudes of these quantities for individual triangles averaged for all layers over the period of investigation have a marked areal dependence as shown in Figs. 16 and 17. Note that at an area of 4^o lat^2, Landers' value (1.5×10^{-5}) is somewhat larger than we obtained while Ballif's value (0.5×10^{-5}) is smaller.

Panofsky (1951) has explained the relationship of divergence to area as follows. Consider that the mean divergence ($\overline{D}$) in a region of horizontal dimension L is given by $(\Delta u + \Delta v)/L$. The spatial difference $(\Delta u + \Delta v)$ tends to be quasi-constant, with a value on the order of a meter per second. Therefore, $\overline{D} \approx$ (constant/L) where L is measured in meters. A curve having typical divergence (1.1×10^{-5} sec^{-1} at an area of 3^o lat^2) and that follows Panofsky's relation is shown by the dashed line in Fig. 16. This curve may also be obtained by a somewhat different argument. From Gauss' theorem,

$$\oint \nabla \cdot \mathbf{V} \, da = \oint v_n ds \quad \text{or} \quad \overline{D} = A^{-1} \oint v_n ds \qquad (17)$$

where v_n is the component of wind normal to the boundary (s) of the area (A). For convenience of argument, let us consider circular areas so that $\overline{D} = \overline{v}_n\ (2\pi r)/\pi r^2 = 2\overline{v}_n/r$. For the divergence specified above, this gives an average normal component ($\overline{v}_n$) of 0.6 m sec^{-1}. If we consider larger areas, we would expect that $\overline{v}_n$ would tend to remain constant or to decrease in magnitude. If we assume that $\overline{v}_n$ remains constant as area increases, we again obtain the dashed curve. However, if the data points of Fig. 16 are fitted by eye with a straight line, we note that this line gives a slower decrease of $\overline{D}$ with area than the dashed curve. If the line describes the true situation, then the average normal component increases as area increases, which seems implausible. With regard to vorticity,

the same type of arguments can be used. From Stokes' theorem

$$\int_A \mathbf{k} \cdot \nabla \times \mathbf{V} \, da = \int_A \zeta \, da = \oint v_t ds \tag{18}$$

where ζ is relative vorticity and v_t is the wind component tangent to the boundary. Again considering circular areas, $\bar{\zeta} = \bar{v}_t \, (2\pi r)/\pi r^2 = 2\bar{v}_t/r$. For a value of vorticity of 2.2×10^{-5} at an area of 3° lat^2, this gives an average tangential component of 1.2 m sec^{-1}. As area increased we would expect $\bar{v}_t$ to remain constant (dashed curve in Fig. 17) or to decrease. The points in Fig. 17 appear to follow a curve of this shape, except for the last few points, which give values of vorticity that appear too large. Since computed values of divergence and also of vorticity at large areas do not decrease in magnitude in the expected manner, we are lead to suspect that observational errors or departures from assumptions of the computational methods are influencing the results. It is not possible to give a definitive discussion of these points; however, two aspects can be mentioned. Errors of observation tend to produce random noise that is largest for small areas. On the other hand, for large areas (and long distances between observations), the mean wind along a line connecting two observations may not be equal to the arithmetic average of the two observed values, as assumed in the computational procedure. These departures from assumed linearity of wind variations between points are a second source of random noise in the computed values. Such noise probably accounts for most of the discrepancy between the line and dashed curve in Fig. 16. For areas larger than 30-40° lat^2, the errors in divergence (probably indicated roughly by the vertical distance between the line and dashed curve) are evidently of the same magnitude as the divergence indicated by the dashed curve. Thus, kinematical computations of divergence would be of doubtful significance when the spacing between observations exceeds 6 or 7° lat. Since vorticity has larger magnitudes, somewhat larger spacing may be permissible in its computation. Finally, it should be noted that due to the close relationship between vorticity and stream function, the same requirement of data density applies to both quantities.

Fortunately, the two principal undesirable aspects of the kinematical calculations of divergence, vorticity, etc., can be reduced considerably by an objective analysis procedure of the type used in the extratropics to obtain values of meteorological quantities at grid points. Such a procedure can be designed to average several adjacent values, thus tending to eliminate effects of observational errors and non-linear wind variations. At the same time, the procedure can normalize values computed for small areas to a larger area, thus reducing the areal effect on the values.

Therefore, in future work, it is expected that smoothed, grid-point values of the computed quantities will be utilized. The sacrifice that must be made is a loss of detail. At the risk of belaboring the obvious, we mention that if one wishes to obtain accurate analyses of sub-synoptic or meso-scale phenomena, observational data must be obtained at rather small space and time intervals, regardless of whether objective or subjective analysis techniques are used.

Vertical motions (computed from the divergence by means of the continuity equation) are, of course, an important quantity due to their relationship to clouds, precipitation, and regions of moist and dry air (as mentioned in Sec. II-B). We wish to obtain values of w that correlate well with observed processes since such values would presumably be useful in forecasting. Therefore, the fields of vertical motion computed for the period of investigation were compared with surface weather and with humidity. In general, the agreement appeared to be relatively good, as has been found by Rex (1958). Typical examples of humidity and vertical motion are shown in Figs. 4 and 15. The association of downward motion with low humidity and upward motion with high humidity is quite good on 5 May. Comparison with clouds on 8 May is considerably hampered because cloud observations made on islands or at coast lines may not be representative of synoptic scale patterns. Further comparisons should therefore be based upon satellite cloud observations (from both television and radiation sensors), since these show the areal extent of clouds and also give indications of cloud heights and thicknesses.

A summary of the average magnitudes of vertical motion at various altitudes is given in the last column in Table IV. In general, the magnitudes apparently increase upwards, partly due to the larger errors at higher altitudes. If we use estimates of these errors given by Endlich and Clark (1963), the magnitude of vertical motions is reduced very slightly in Layer II and to about 10 cm sec^{-1} in Layer V. Thus, the main increase is apparently real, as mentioned previously by Rex.

At present, the kinematical values of vertical motion are not considered by the writers to be of demonstrable utility. Further work is needed to combine them with vertical motion determined by other techniques--in particular, from Eq. (3), and moreover to investigate their inclusion in numerical forecasting models.

E. SUMMARY OF THE INVESTIGATION

It is believed that the work that has been completed indicates strongly that objective techniques of data handling and analysis are as applicable in tropical regions as they have proved to be in extratropical meteorology. In many cases, actual procedures for the tropics must differ from accepted procedures in mid-latitudes for reasons discussed earlier. However, in the writers' opinion, numerical methods of forecasting have as much potential in the tropics as elsewhere. The main need is for further research in this rather neglected field. It is thought that the techniques of analysis discussed in this report are suitable tools for further investigations of tropical weather phenomena and for the further development of computer methods of analysis and forecasting. It is probable that the technique of stream function computation (Sec. II-C) is also applicable to stratospheric analysis (where heights are less reliable than winds), and, in fact to the analysis of wind data anywhere on the globe.

There are numerous subjects that require further research. One topic of importance is a comparison of the stream function as computed herein with a solution based on Poisson's equation, both using the same data. A second topic is to explore the feasibility of kinematical advection (Sec. II-B) as a forecasting method and also as a basis for judging

dynamical forecasts. Another topic is the computation of a height field compatible with the stream function field. Such a computation might be made by use of the balance equation, solving for height from the known values of ψ. It is also possible to attempt to solve for the velocity potential χ by use of a modified form of the stream function procedure. An approach for doing this has been described at the end of Sec. II-C. Further investigation is also needed concerning the computation of vertical motions, and the relationships of vertical motions and tropical storms where the latter are shown by satellite and conventional data. The final topic of importance is the crux of the problem, namely the development of dynamical prediction models applicable to tropical forecasting. To a large extent, the experience of extratropical meteorologists can be used as a guide; nevertheless, a series of successively more complex experiments carried out over a period of several years will be required to reach this goal.

PERSONNEL

Name	Approximate Hours Devoted to Project
M. G. H. Ligda, Project Supervisor	30
R. M. Endlich, Project Leader	830
R. L. Mancuso, Research Meteorologist	1250
J. R. Clark, Research Meteorologist	95
W. Viezee, Research Meteorologist	20
J. Weaver, Meteorological Aid	310
B. Crowell, Meteorological Aid	100

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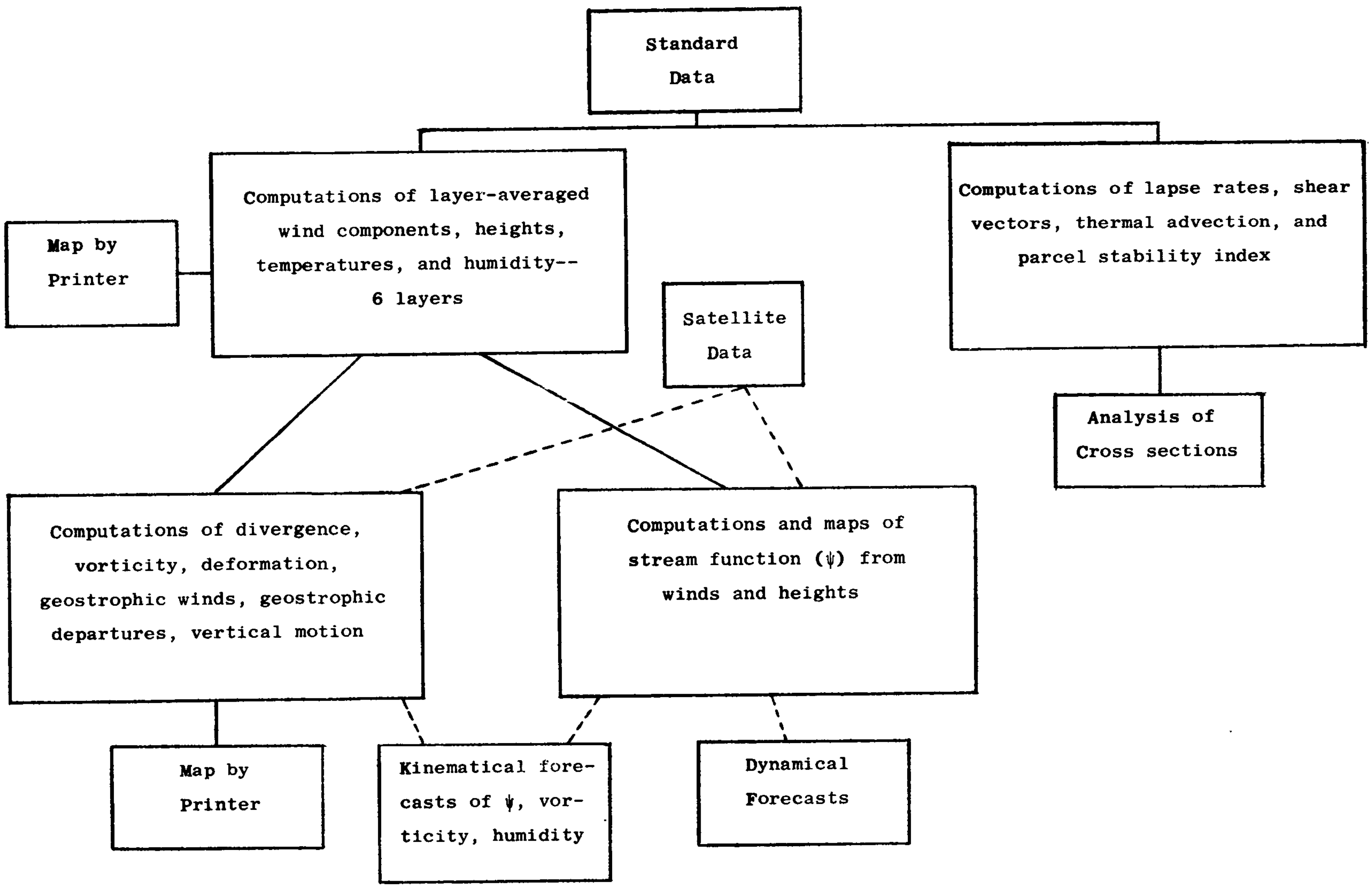


Figure 1 **Schematic Diagram of Objective Analysis and Forecasting in the Tropics**

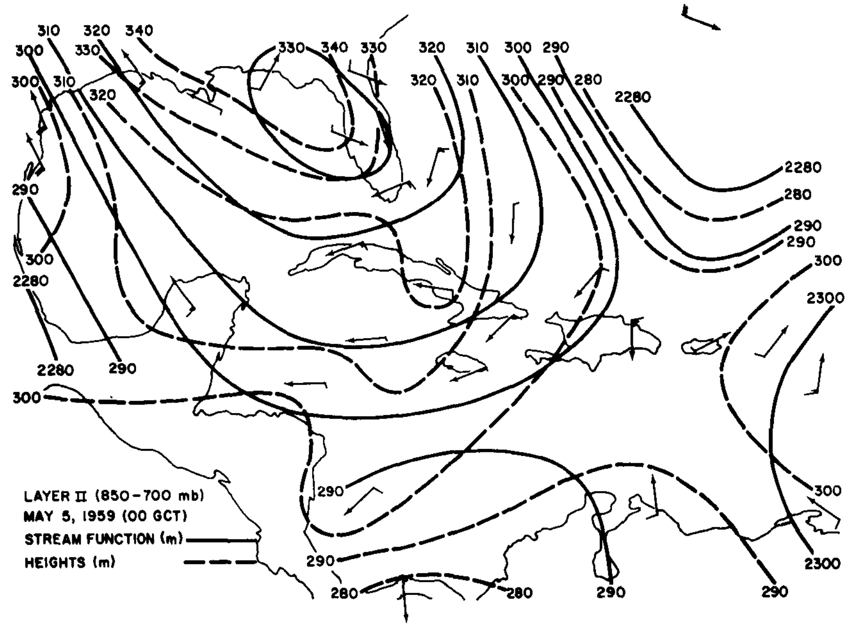


FIG. 2 OBSERVED HEIGHTS AND VALUES OF STREAM FUNCTION IN LAYER II, 00GCT ON 5 MAY 1959

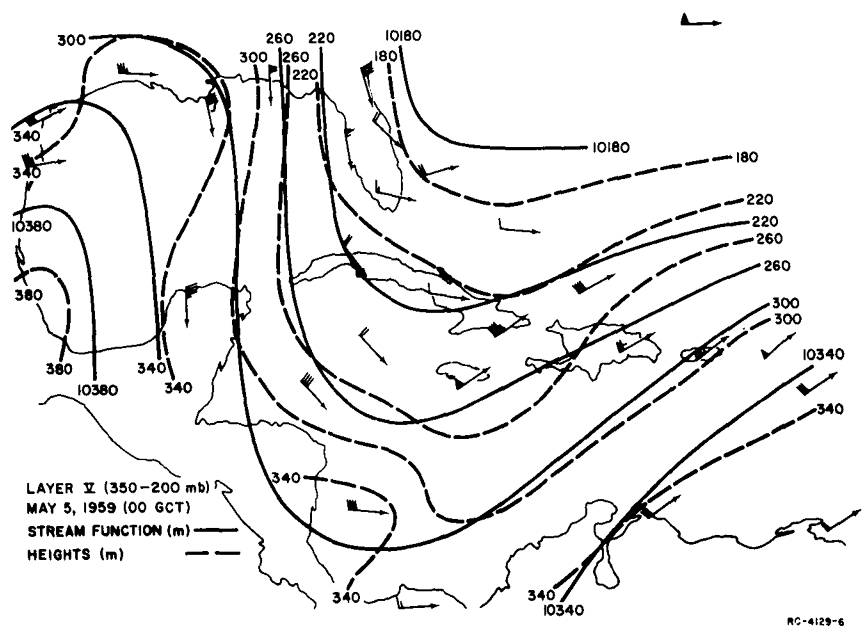


FIG. 3 OBSERVED HEIGHTS AND VALUES OF STREAM FUNCTION IN LAYER V.

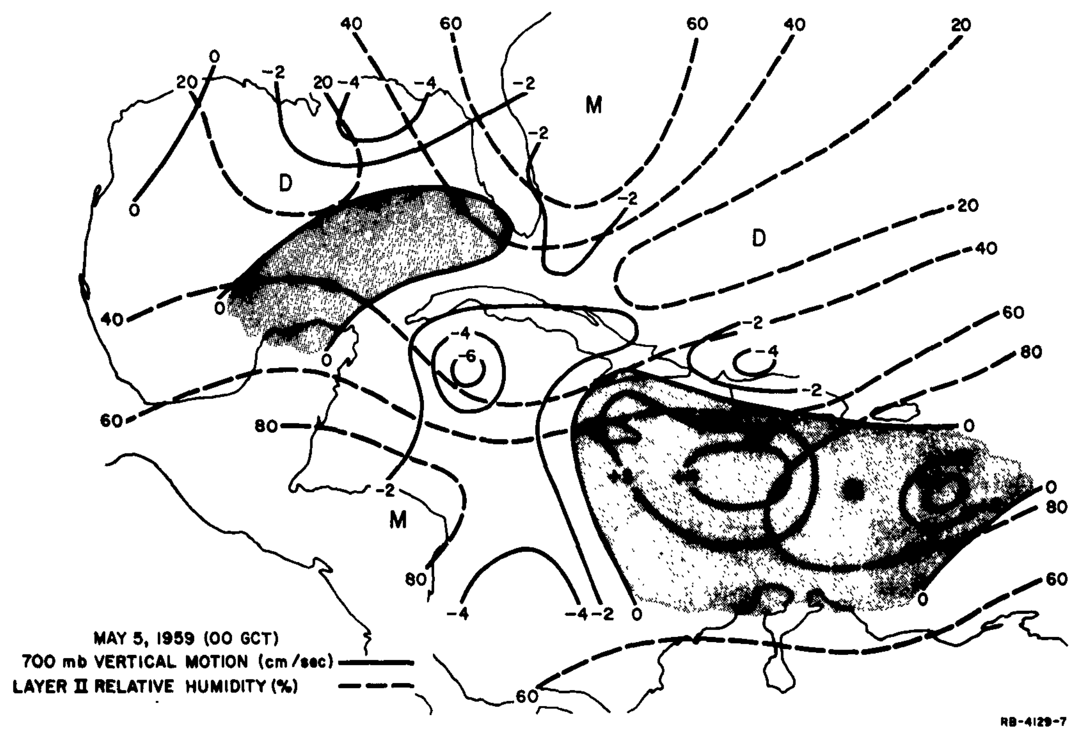


FIG. 4 VERTICAL MOTION AT 700 mb, AND RELATIVE HUMIDITY IN LAYER II. SHADED REGION HAS UPWARD MOTION.

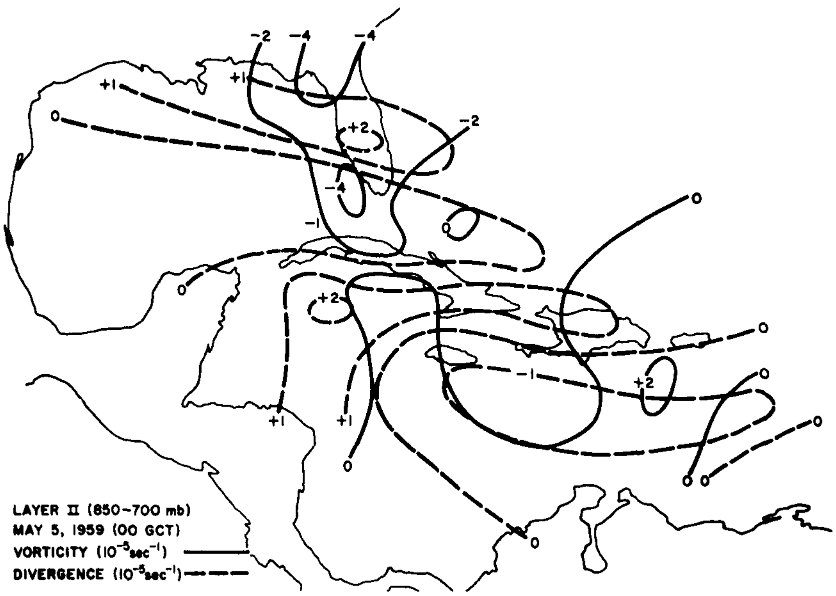


FIG. 5 RELATIVE VORTICITY AND DIVERGENCE IN LAYER II

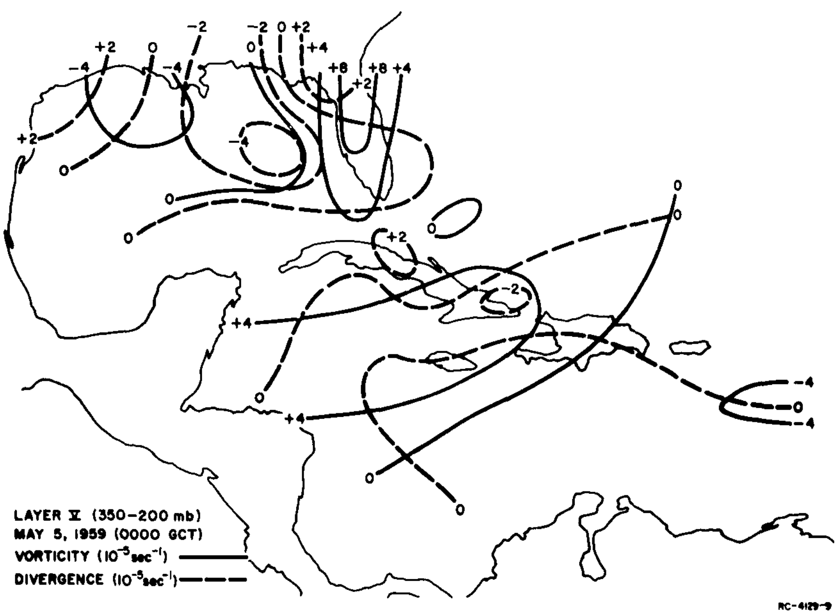


FIG. 6 RELATIVE VORTICITY AND DIVERGENCE IN LAYER V

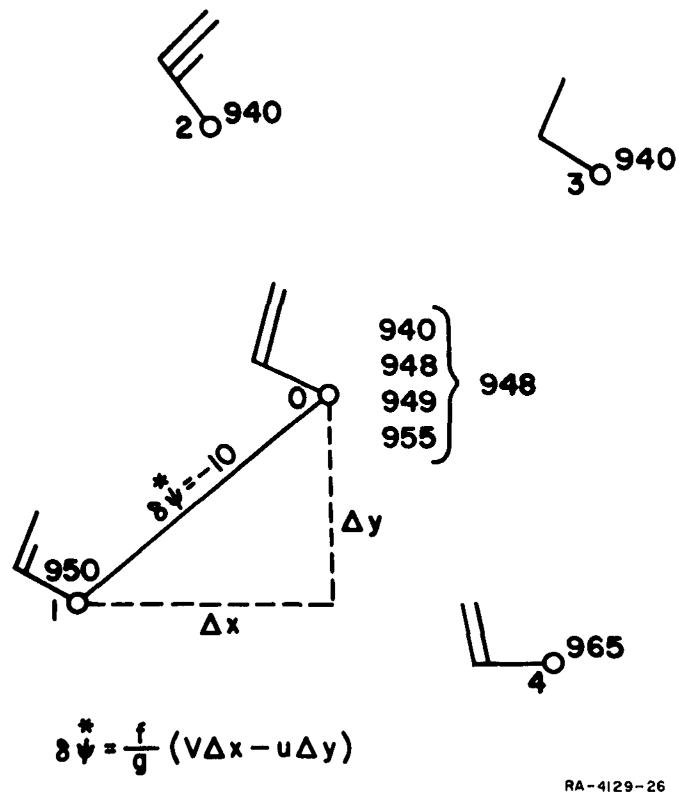


FIG. 7 HYPOTHETICAL EXAMPLE OF THE STREAM FUNCTION COMPUTATION AT STATION 0

OBJECTIVE STREAM FUNCTION PROGRAM

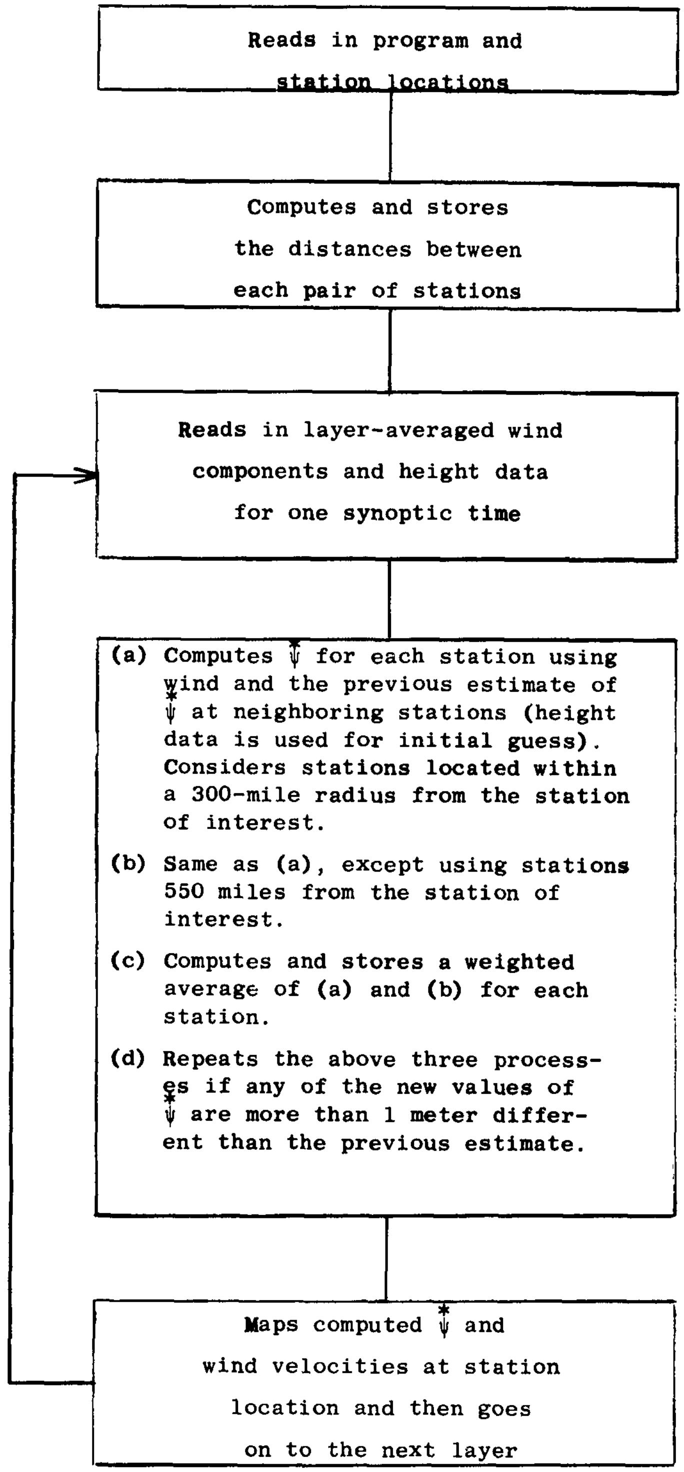


Figure 8 **FLOW DIAGRAM OF COMPUTATIONS--STREAM FUNCTION**

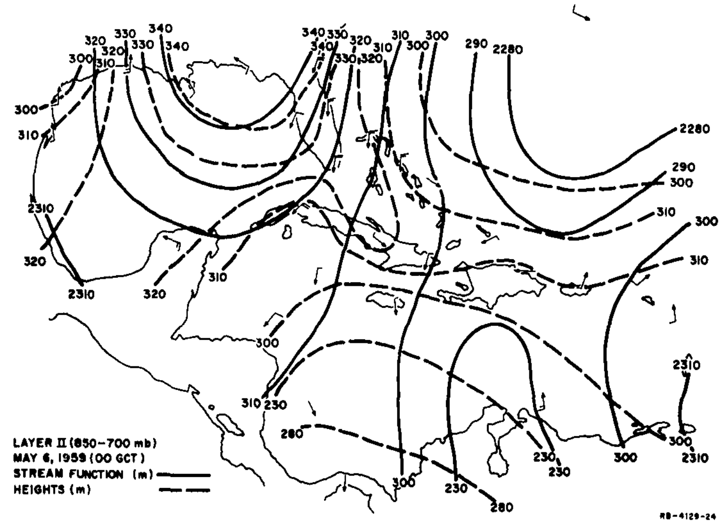


FIG. 9 OBSERVED HEIGHTS AND VALUES OF STREAM FUNCTION IN LAYER II, OOGCT ON 6 MAY 1959

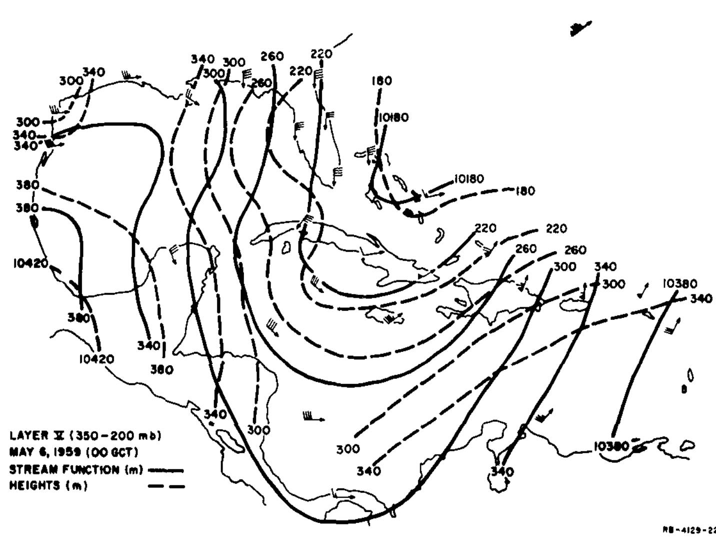


FIG. 10 OBSERVED HEIGHTS AND VALUES OF STREAM FUNCTION IN LAYER V

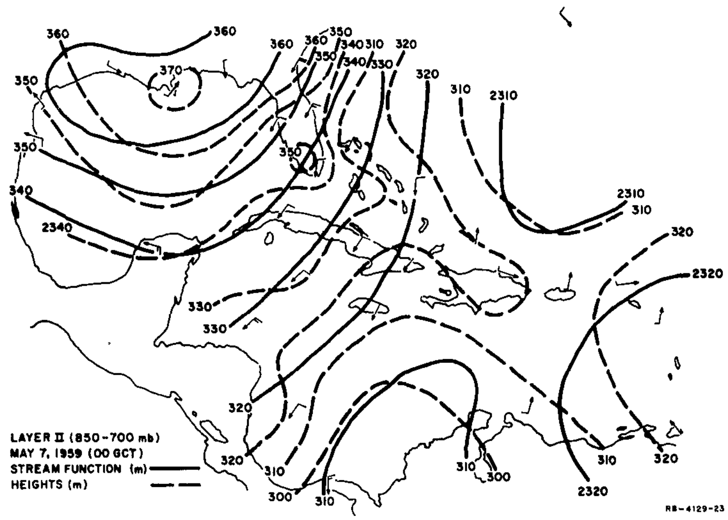


FIG. 11 OBSERVED HEIGHTS AND VALUES OF STREAM FUNCTION IN LAYER II, OOGCT ON 7 MAY 1959

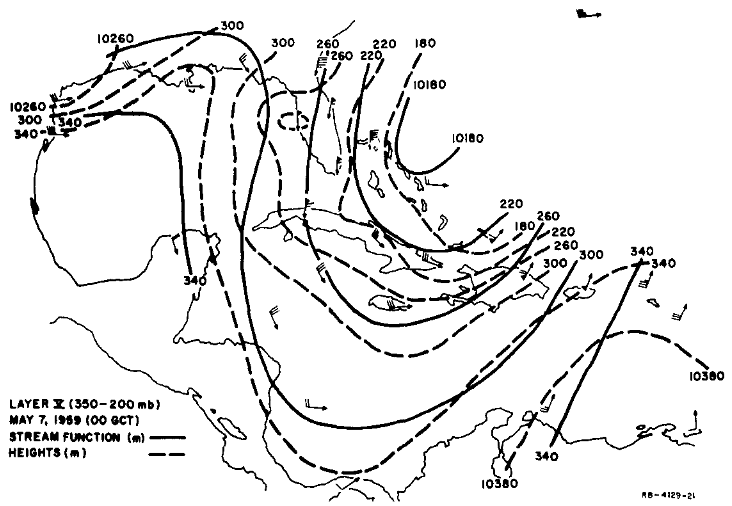
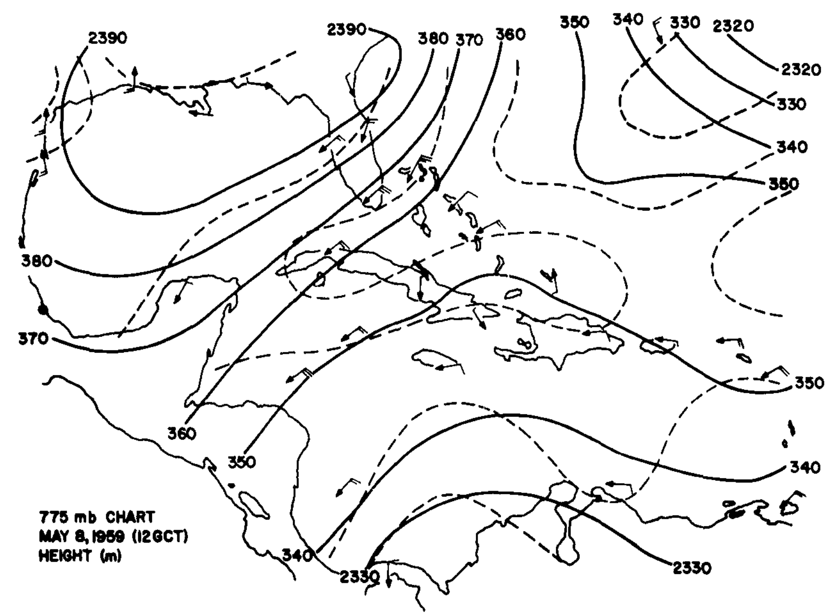


FIG. 12 OBSERVED HEIGHTS AND VALUES OF STREAM FUNCTION IN LAYER V



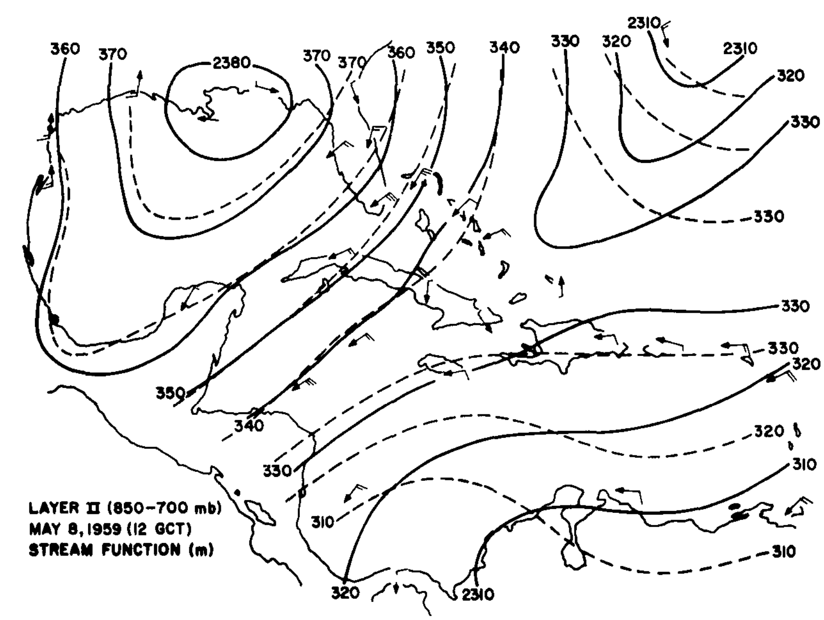


FIG. 13 OBSERVED HEIGHTS, SMOOTHED HEIGHTS, AND STREAM FUNCTION IN LAYER II 12GCT ON 8 MAY 1959. UPPER: DASHED LINES ARE DRAWN TO OBSERVED HEIGHTS. SOLID LINES ARE SMOOTHED HEIGHT ANALYSIS MADE BY PORTIG. LOWER: DASHED LINES ARE DRAWN TO STREAM FUNCTION VALUES AFTER ONE SCAN. SOLID LINES ARE FINAL VALUES OF STREAM FUNCTION.

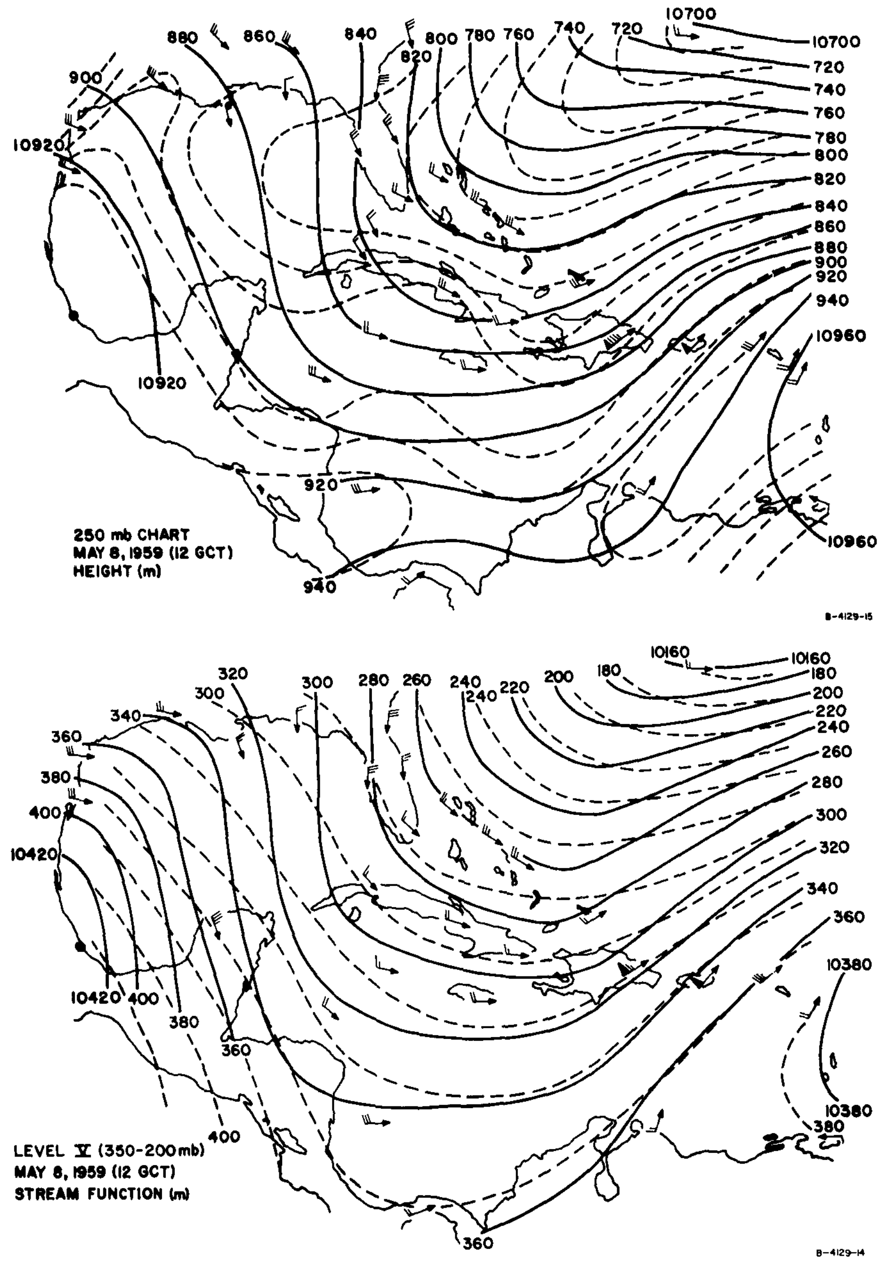


FIG. 14 OBSERVED HEIGHTS, SMOOTHED HEIGHTS, AND STREAM FUNCTION IN LAYER V. UPPER: DASHED LINES ARE DRAWN TO OBSERVED HEIGHTS. SOLID LINES ARE SMOOTHED HEIGHT ANALYSIS MADE BY PORTIG. LOWER: DASHED LINES ARE DRAWN TO STREAM FUNCTION VALUES AFTER TWO SCANS. SOLID LINES ARE FINAL VALUES OF STREAM FUNCTION.

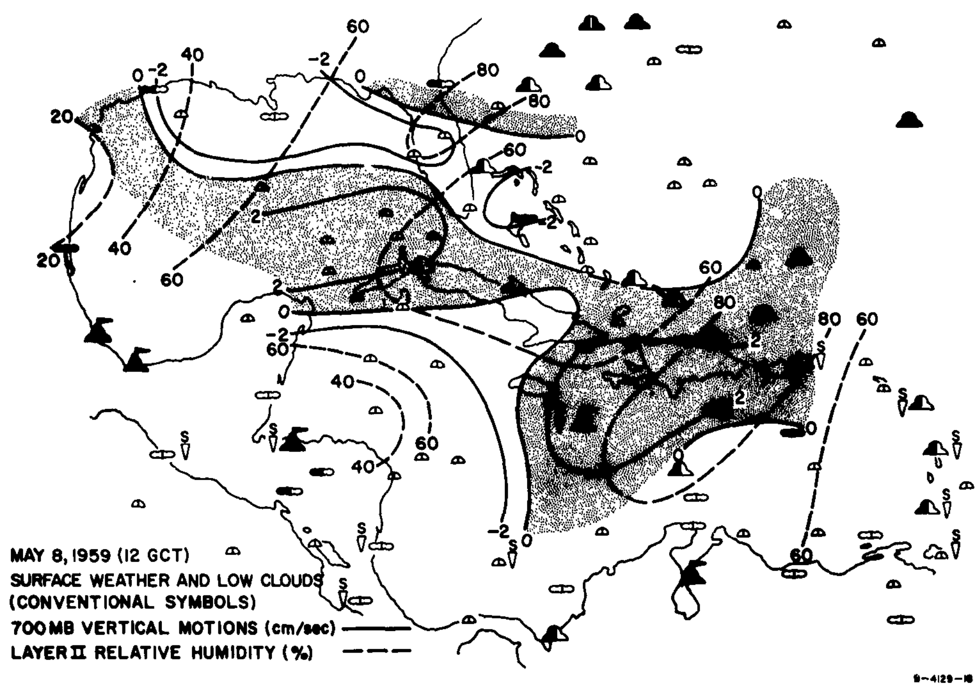


FIG. 15 VERTICAL MOTION AT 700 mb, RELATIVE HUMIDITY IN LAYER II, AND REPORTED CLOUDS AND WEATHER

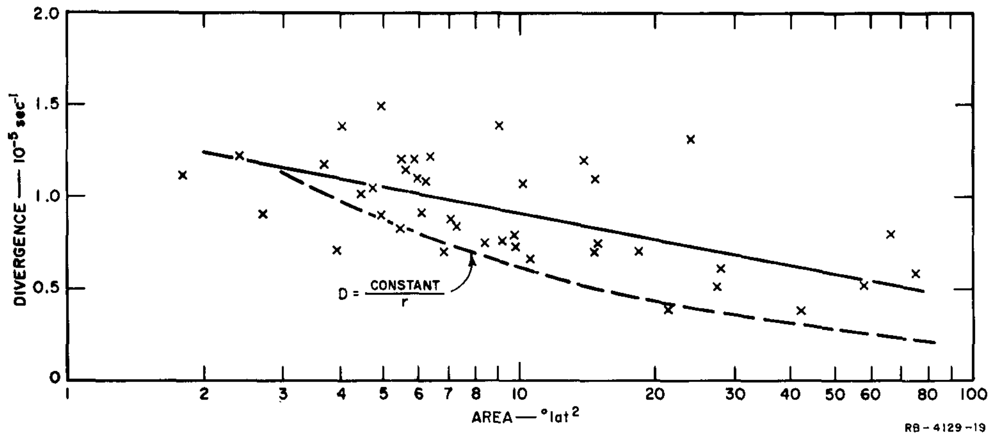


FIG. 16 AVERAGE MAGNITUDE OF DIVERGENCE AS A FUNCTION OF THE AREA OF THE TRIANGLE USED IN THE COMPUTATION (Caribbean data for 5-8 May 1959)

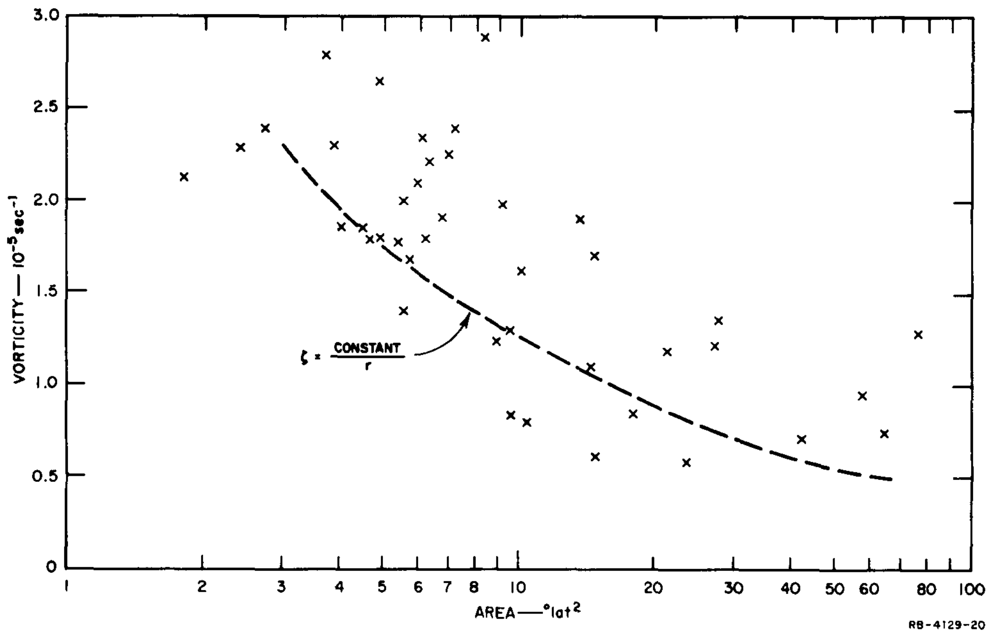


FIG. 17 AVERAGE MAGNITUDE OF VORTICITY AS A FUNCTION OF THE AREA OF THE TRIANGLE USED IN THE COMPUTATION

APPENDIX A

We wish to obtain an equation which relates the changes in relative humidity experienced by an air parcel to vertical motions. The development is similar to that given by Smagorinsky (1960). We begin with the Clausius-Clapeyron equation:

$$de_s/dT = (\epsilon L e_s / R_d T^2)$$

where e_s is the saturation vapor pressure, T is temperature, ϵ equals 0.622, L is the latent heat of condensation, and R_d is the gas constant for dry air. Under the assumption of dry adiabatic motion, the first law of thermodynamics is:

$$c_p(dT/dt) = (\rho)^{-1}(dp/dt) = (\rho)^{-1}\omega \quad .$$

Also for dry adiabatic motion, specific humidity (q) is conserved, *i.e.,* $(dq/dt) = 0$ where $q = (\epsilon e/p)$, e being the vapor pressure. The relative humidity is:

$$h = e/e_s = q/q_s \quad .$$

On differentiating with respect to time we obtain

$$dh/dt = (q_s)^{-1}(dq/dt) - (q/q_s^2)(dq_s/dt)$$

$$= -(h/q_s)(dq_s/dt) \quad .$$

Since $q_s = (\epsilon e_s/p)$, we obtain

$$dh/dt = -(\epsilon h/q_s)[(p)^{-1}(de_s/dT)(dT/dt) - (p)^{-2}(\omega e_s)]$$

$$= -(\omega h/p)[(\epsilon L/c_p T) - 1]$$

This form relates dh/dt and ω. To obtain vertical motions in terms of w, let $\omega = -\rho g w$ so that:

$$dh/dt = (wgh/RT)[(\epsilon L/c_p T) - 1] \quad \text{or}$$

$$w = (dh/dt)(T^2/h)[(c_p R/g)(\epsilon L - c_p T)^{-1}] \quad .$$

For w in cm $\sec^{-1}$, h in percent, T in °K, and t in seconds, the quantity in brackets has an average value of approximately 2.3. Thus,

$$w = 2.3(dh/dt)(T^2/h)$$

It should be borne in mind that dh/dt contains a term $w\ \partial h/\partial z$. To solve for w we can write the previous equation as:

$$w - [w(\partial h/\partial z)(T^2/h)(2.3)] = (\partial h/\partial t + \mathbb{V}_H \cdot \nabla h)(T^2/h)(2.3) \quad .$$

The terms $\partial h/\partial t$ and $\mathbb{V}_H \cdot \nabla h$ can be evaluated by standard methods. As a first approximation in computing w, it appears that the second term on the left hand side can be neglected. Then the equation can be iterated including this term to obtain the correct value of w.

APPENDIX B

To obtain formulas for kinematical advection of meteorological quantities which are quasi-conservative, we follow the outline of Petterssen (1940). In fixed coordinates, the particle derivative of a scalar quantity is

$$\frac{d\alpha}{dt} = \frac{\partial \alpha}{\partial t} + \mathbb{V} \cdot \nabla \alpha \tag{1}$$

Similarly, in a system of moving coordinates the parcel derivative is

$$\frac{d\alpha}{dt} = \frac{\delta \alpha}{\delta t} + \mathbb{V}' \cdot \nabla \alpha \tag{2}$$

where $\delta\alpha/\delta t$ is the local change with respect to the moving coordinates and $\mathbb{V}'$ is the wind velocity relative to the moving coordinates. On equating equations (1) and (2)

$$\frac{\delta \alpha}{\delta t} = \frac{\partial \alpha}{\partial t} + \mathbf{C} \cdot \nabla \alpha \tag{3}$$

where $\mathbf{C} = \mathbb{V} - \mathbb{V}'$ is the velocity of the moving coordinate system (with respect to the fixed coordinates). If α is a conservative quantity, $\delta\alpha/\delta t = 0$. Petterssen discusses cases where $\alpha = p$ (isobar), $\alpha = \partial p/\partial t$ (isallobar), etc. For quasi-conservative quantities, $\delta\alpha/\delta t$ is considerably smaller than the other terms in equation (3) so that

$$\frac{\partial \alpha}{\partial t} = -\mathbb{C} \cdot \nabla \alpha \qquad (4)$$

We wish to use this equation to obtain $\mathbb{C}$ (the speed of movement of a quantity of interest) from a knowledge of $\partial\alpha/\partial t$ (estimated at a point over the interval between observations) and $\nabla\alpha$ (given at a particular time or succession of times) over a weather chart. If we consider a point on a curve α = constant, then $\nabla\alpha$ is perpendicular to this curve and the velocity desired is that along $\nabla\alpha$ so that

$$\frac{\partial \alpha}{\partial t} = -C \frac{\partial \alpha}{\partial n} \qquad (5)$$

where C is the speed of movement of the curve at the point of interest and n is along $\nabla\alpha$. Equation (4) may also be written as

$$\frac{\partial \alpha}{\partial t} = -\left(u_c \frac{\partial \alpha}{\partial x} + v_c \frac{\partial x}{\partial y}\right) \qquad (6)$$

The relationships between these various quantities are shown in Fig. B-1. Equation (6) can be written as

$$\begin{aligned} \frac{\partial \alpha}{\partial t} &= -u_c\left(\frac{\partial \alpha}{\partial x}\right) - u_c \tan\theta\left(\frac{\partial \alpha}{\partial y}\right) \\ &= -u_c\left[\left(\frac{\partial \alpha}{\partial x}\right) + \left(\frac{\partial \alpha}{\partial y}\right)\left(\frac{\partial \alpha}{\partial x}\right)^{-1}\left(\frac{\partial \alpha}{\partial y}\right)\right] \\ &= -u_c\left[\frac{\left(\frac{\partial \alpha}{\partial x}\right)^2 + \left(\frac{\partial \alpha}{\partial y}\right)^2}{\left(\frac{\partial \alpha}{\partial x}\right)}\right] \end{aligned}$$

or

$$u_c = \frac{-\left(\frac{\partial\alpha}{\partial t}\right)\left(\frac{\partial\alpha}{\partial x}\right)}{\left(\frac{\partial\alpha}{\partial x}\right)^2 + \left(\frac{\partial\alpha}{\partial y}\right)^2} \tag{7a}$$

Similarly,

$$v_c = \frac{-\left(\frac{\partial\alpha}{\partial t}\right)\left(\frac{\partial\alpha}{\partial y}\right)}{\left(\frac{\partial\alpha}{\partial x}\right)^2 + \left(\frac{\partial\alpha}{\partial y}\right)^2} \tag{7b}$$

It is believed that these equations can be used to obtain forecasts by an iterative method. If present and past values are denoted by the subscripts 0 and −1, we can determine the mid-values $(\partial\alpha/\partial t)_{-\frac{1}{2}}$, $(\nabla\alpha)_{-\frac{1}{2}}$, and $(\mathbf{C})_{-\frac{1}{2}}$. Then from $(\mathbf{C})_{-\frac{1}{2}}$ and $(\nabla\alpha)_0$ we obtain a first estimate of $(\partial\alpha/\partial t)_{\frac{1}{2}}$. Using the formula

$$\alpha_1 = \alpha_0 + (\Delta t)(\partial\alpha/\partial t)_{\frac{1}{2}}$$

we obtain the first estimate of the future value of α, *i.e.*, α_1. The procedure may be repeated several times to improve the estimate of α_1.

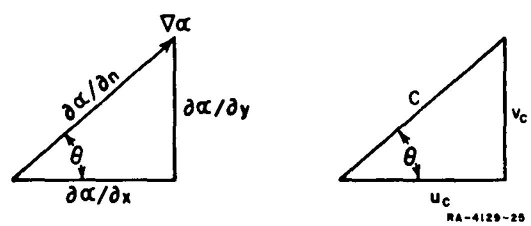


FIG. B-1 RELATIONSHIPS BETWEEN $\nabla\alpha$, $\mathbf{C}$, AND θ

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Computations were also made of divergence, vorticity, deformation, and vertical motion in each of the layers. The average magnitude of divergence (for an average area of approximately 10° lat²) was 0.9×10^{-6} sec^{-1} while the magnitude of relative vorticity was approximately twice as large. The dependence of the magnitudes upon scale was also investigated. An equation is given for computing vertical motion from the change of relative humidity experienced by an air parcel. This equation is believed to be appropriate for use in the tropics, where large spatial and time changes in humidity are observed. An equation is also given for the kinematical advection of quantities such as stream function, humidity, and vorticity. Requirements for further research concerning numerical techniques are summarized. It is concluded that objective techniques of analysis and forecasting have as much potential value in the tropics as elsewhere.	relative humidity subjective analyses computer program stream function observed winds observed heights vorticity divergence vertical motion UNCLASSIFIED
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